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nocturnal air temperature on a forested mountain slope

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Rocky Mountain Forest and Range Experiment Station
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[1]Central headquarters maintained at Fort Collins, in cooperation with Colorado State University.

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James D. Bergen

In mountainous terrain, the spatial pattern of nighttime temperatures measured in standard screens should largely reflect topography. Unfortunately, towns and permanent weather stations are commonly located in canyons and river valleys, while the ecologist and forest-protection officer are usually concerned with the temperatures found on the forested slopes above such stations; they are also less concerned with the temperature at a standard screen height than with the temperature at some level in the forest canopy or at the surface of the slope.

The common assumption that air temperatures vary with elevation on a slope in the same way that air temperatures vary with altitude in the nearby free atmosphere is plausible only when the net radiation at the slope surface approaches zero. This situation is possible on days or nights with considerable low cloud cover and in early evening for most days, but not on those still, clear nights on which minimum temperatures are likely to occur. For such nights the movement of air and the pattern of surface isotherms follow the details of the topography in a complex way. The cold air layer generated by radiation cooling is shallow, with a depth on the order of tens of meters. Such layers may not extend far above the forest canopy, and small topographic features such as ridges or gullies interact directly with the structure of the canopy to determine the vertical temperature profile at any position along the slope.

The data to be discussed in this Paper were collected during an exploratory study of the joint effect of these factors on the nocturnal temperature of a highly divergent, heavily forested slope typical of much of the Colorado Rockies.

Study Site

A slope of simple geometry was selected for the study; the elevation contours are shown in figure 1. The forest cover is a mixed stand of Engelmann spruce (*Picea engelmannii* Parry) and subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.), which attain a maximum height of about 18 meters (m.). The hill rises to a maximum height of about 200 m. above the Fraser River valley near Fraser, Colorado, and faces east with a relatively open horizon.

Measurement stations established on the slope (fig. 1) were chosen to approximate points along a streamline for the cold air movement off the slope; the assumption was that such movement is normal to the elevation contours. When a number of adjacent streamlines are estimated by this assumption (fig. 1A), a feature is emphasized which may be noted only fleetingly from the contour maps, and which would not be apparent from a two-dimensional cross section (fig. 1B). This feature is a center of divergence, or more rigorously, the end of a line of divergence labeled (C). Point (C) may be shown to be a region where the motion of air in response to temperature differences between the air on the slope and that over the valley must be primarily normal to the slope. This point may, therefore, be regarded as a probable intake zone for the movement of cold air off the hillside.

Instruments

The stations instrumented and the levels at which the instruments were placed are shown for each year (table 1). The instrument profiles were

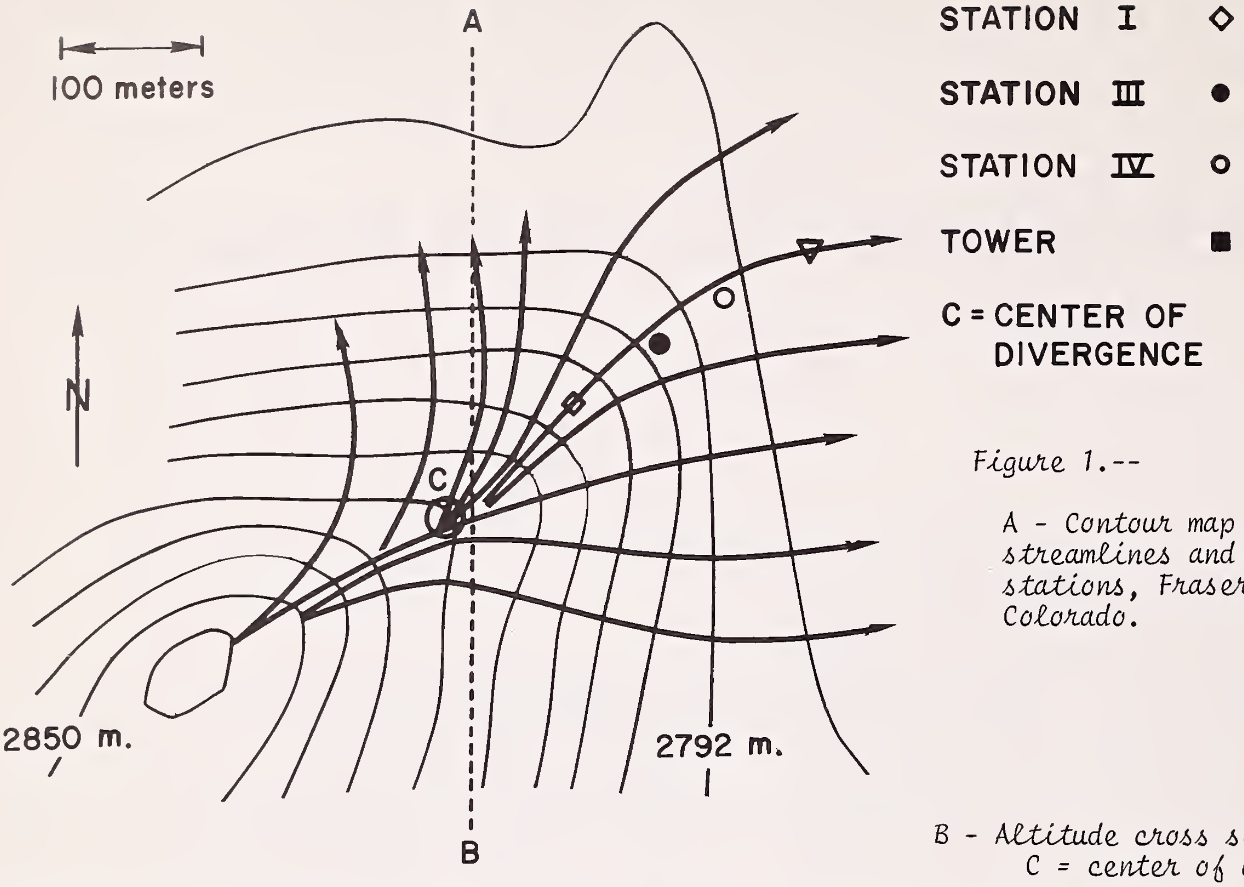
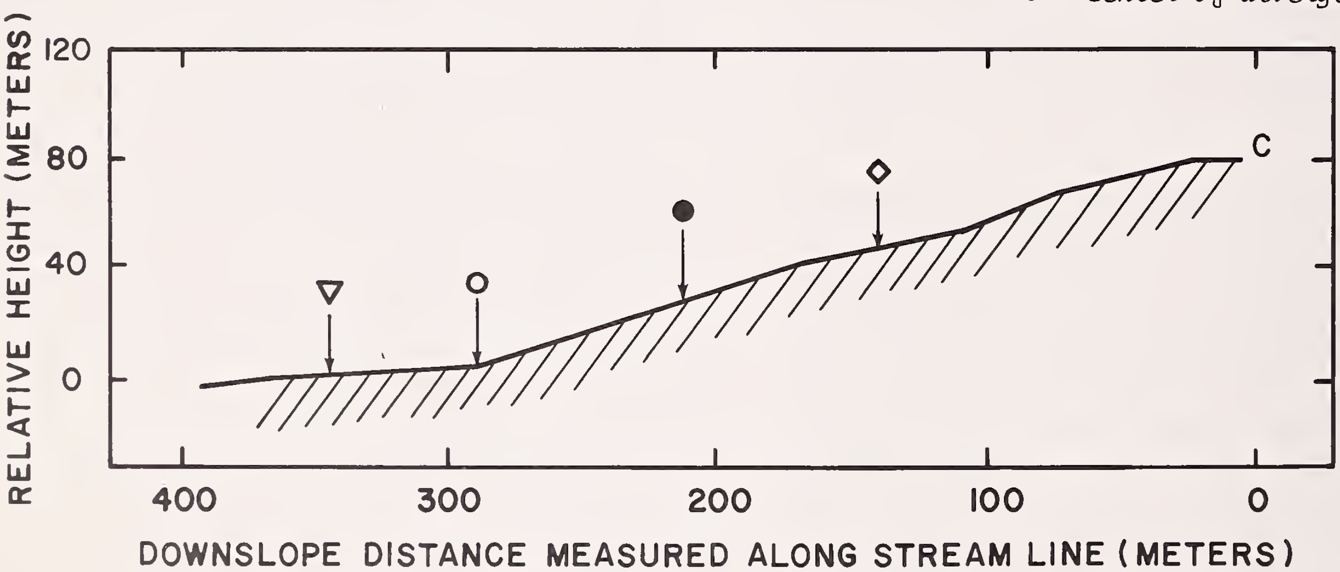


Figure 1.--

A - Contour map with estimated streamlines and measurement stations, Fraser River valley, Colorado.

B - Altitude cross section of transect. C = center of divergence.



placed approximately between trees, with each profile having at least 5 m. of clear fetch up the slope.

Air temperatures were measured with precision bead thermistors, mounted on 1-m.-long arms that extended across slope from the nearest tree stem and well away from foliage. The nominal resistance of these beads is 10 KΩ, high enough so that thermal effects even along a considerable length of cable were negligible. The thermal time constant of these beads is about 2 seconds, but smoothing was provided by radiation shields which enclosed each bead. For the first 2 years, these were of folded aluminum. In 1966, a half-closed styrofoam cup was used as a shield. The time constant with moderate ventilation was on the order of a minute.

At the tower near the slope base (T in fig. 1), shielded thermocouples were used at the 0.5-, 2.5-, 5.0- and 15.0-m. levels. In 1962 and 1964, temperatures at the 0.3-m. level on the slope were measured with bimetal thermographs. One of these instruments was also located in a screen at a height of 3 m., about 100 m. to the east of the base tower in 1964 and 1966.

Table 1.--Stations measured and levels on slope at which instruments were placed, 1962, 1964, 1966

Year	Stations measured	Location of instruments								
		Air-temperature levels								Other
		- - - Meters - - -								
1962	I[1]	0.3	1.7	5.0						
	II[1]	.3	1.7	5.0						
	III	.3	1.7	5.0						
	IIIb	.3	1.7	5.0						
	IV[1]	.3	1.7	5.0						
	Tower	1.3	2.5	5.0	10.0	20.0				
1964	I	.7	6.0	7.0	9.0	12.0				Surface pressure
	III	.7	10.0	12.0						Surface pressure
	IV	.7	2.3	4.3	6.0	8.3				Surface pressure
	Tower	1.3	2.5	5.0	10.0					Total radiation
	Screen	3.0								
1966	I	.5	1.8	3.9	14.0					Surface pressure
	III	.4	1.7	3.1	5.9	7.4	8.7	10.2	14.6	
	IV	.3	1.5	3.0						Surface pressure
	Tower	1.3	2.5	5.0	5.9	10.0	11.4	20.0		Total radiation
	Screen	3.0								

[1]Two profile stations 20 m. apart on either side of the transect.

The nominal observation period for all years was 1800 local standard time (LST) to 0600 LST of the following day. Observations were begun only on clear afternoons, and when the 20-m. tower wind was less than 0.5 m. sec.$^{-1}$ at 1800 LST. However, conditions often changed during the night; August 25, 1962, was relatively cloudy by 2000 LST.

In the 1962 observations, the thermistor resistances were measured with a manually operated bridge circuit and a plug board; temperature profiles at stations Ia, IIa, IVa, and Ib, IIb, IVb were read every 20 minutes. The output of the tower thermocouples was recorded automatically at 12-minute intervals. For the 1964 and 1966 observations, all temperatures except at the bimetal thermograph positions were measured and recorded automatically at 12-minute intervals.

Total radiation was measured with a ventilated radiometer whose output was recorded at 12-minute intervals during observation periods. Occasional checks were made for signs of dew or frost on foliage or slope surface. None was observed for any of the nights.

Pressure variations on the slope were recorded on aneroid microbarographs.

Vertical Temperature Profiles

The most detailed vertical temperature profiles in the observation set were those measured at station III, called the "midslope station" below, in 1966. A typical midslope station profile is shown in figure 2. This profile is based on the 2-hour average of air temperature from 2400 LST on the observation night of June 15, 1966. The lowest temperatures and strongest vertical temperature gradients appear at the slope surface, as would be anticipated if the latter functioned as a principal heat sink. The vertical gradient decreases sharply above the first meter of height, to the region from 4 to 6 m., increases, and then vanishes at an inversion at about 7 m. height where the temperature shows a local maximum. Higher temperatures are not attained on the profile below a level of 11 m. height. This layer of air is statically unstable; that is, a relatively warm layer is underlying a colder layer of air.

The entire cooled layer at this station extends to about 30 m., as judged from the temperature drop from the 1800 LST profile and the pressure increment from that time indicated by the micro-

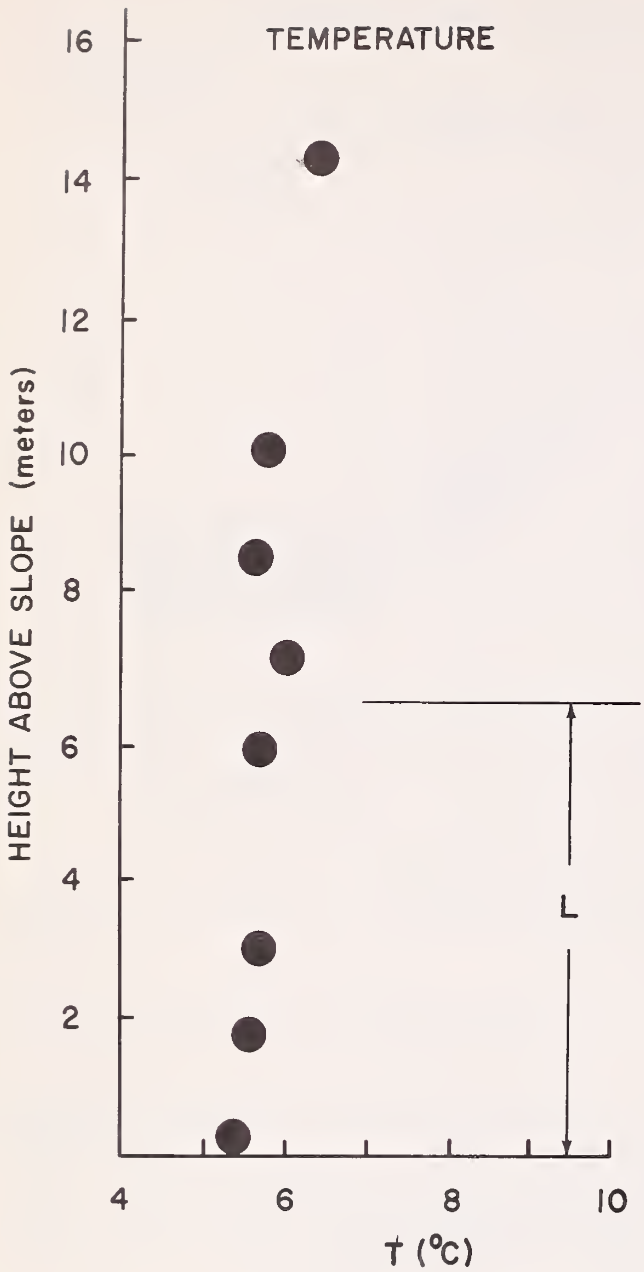


Figure 2.--Midslope station temperature profile, Station III, June 15, 1966, 2400-0200 LST. L = height of first inversion above slope.

barograph trace. The height of the first inversion above the slope is labeled (L) on figure 2, and will be referred to in the subsequent analysis of the temperature data.

The most interesting feature of the profile is obviously the warm layer at the first inversion. This feature occurs in the temperature measurements presented by Geiger (1958). His measurements were vertical profiles of air temperature along a strip cut down the side of a conifer-covered mountainside, and they are in most respects comparable to those described in this paper. His measurements were taken at various positions along a geometrical transect of the mountainside, however, which may not have had any direct relation to the pattern of air movement from the slope. Geiger's measurements indicate an "altitude of maximum temperatures," which separates the vertical temperature profiles into two types. The lower region shows a warm zone below the treetops, quite analogous to that found on figure 2 for the Fraser measurements; the upper region does not.

Variation of Temperature with Downslope Distance

A characteristic temperature which can be readily defined from the vertical temperature profiles of the 1966 observations, and which can be estimated from the observations of both the preceding years (1962 and 1964) is the average temperature ($\overline{T}$) of the air layer below the height (L).

This average was computed on the assumption that (L) was approximately the same for all the nights of observation, with a value of 7 m. at the midslope station III, 10 m. at the tower station, and 8 m. at station I. The value of (L) at the intermediate stations was interpolated from these points. As seen from the relatively small temperature gradients near (L), this average temperature is not particularly sensitive to differences of the order of 1 m. in the assumed value of (L).

The hourly average of this temperature is plotted for each of the 13 nights of observation (fig. 3).

At the beginning of the evening—1800 LST—the temperatures either show a decrease or no variation down the transect. Six of the nights show no variation. A somewhat more irregular pattern is evident for September 17 and 25, 1964, and June 14, 1966. On September 17, a definite temperature maximum occurs on the slope at the vicinity of station III, which persists well into the evening. On September 25, a temperature minimum is evident near station IV, which vanishes near midnight and reappears the following morning. This effect may reflect a relatively uncommon large-scale stratification over the Fraser River valley.

In general, the temperature pattern for the night may be divided into two periods similar to those

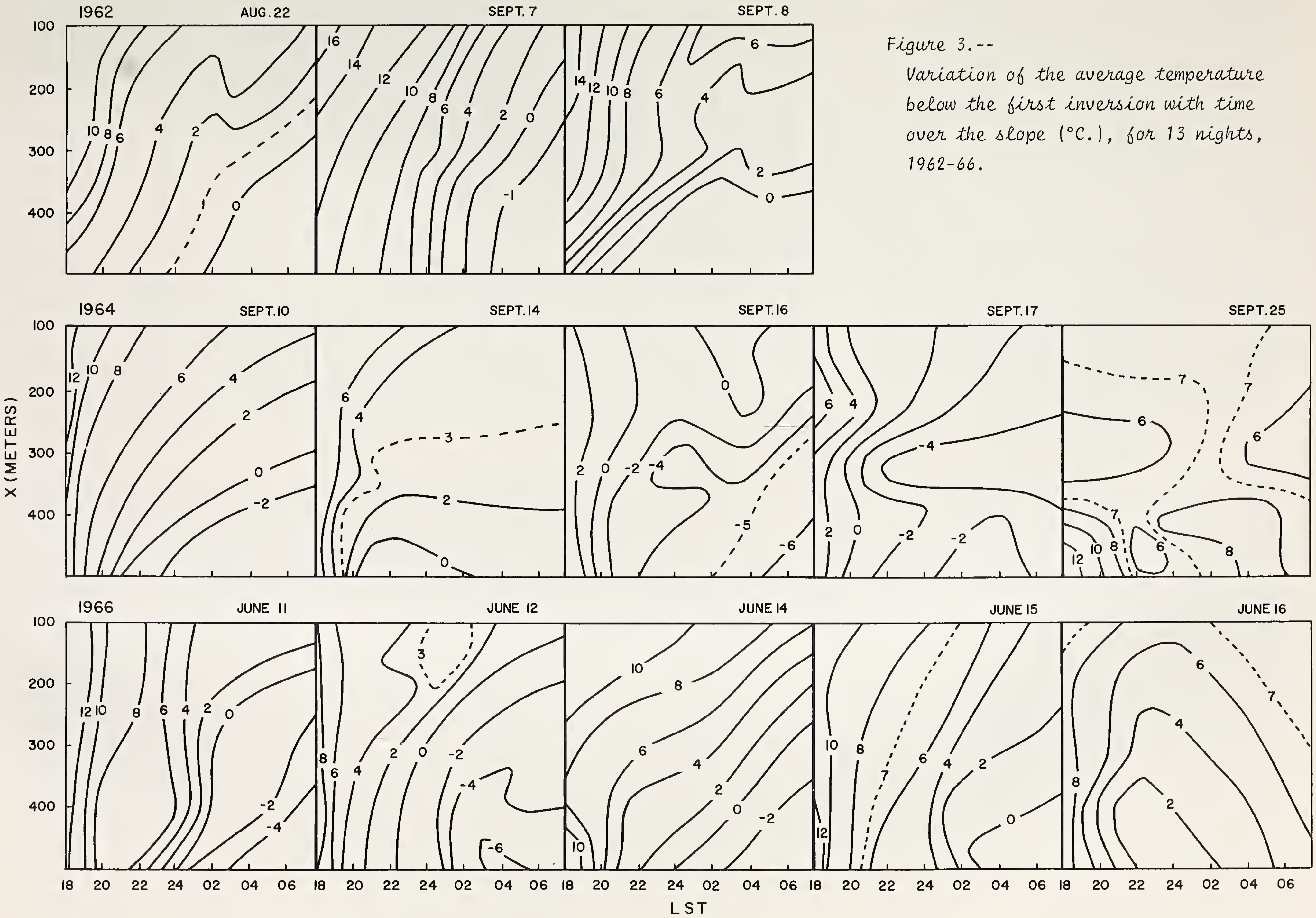


Figure 3.--
Variation of the average temperature below the first inversion with time over the slope (°C.), for 13 nights, 1962-66.

found in the study by Leighton (1954)—a period of uniform cooling over the entire slope with a relatively constant temperature gradient down the slope, and a period of increasing downslope gradient with little temperature change near the crest of the hillside (point C).

On some nights, for instance June 11, 1966, the second phase does not seem to occur. On others, notably June 15, 1966, the first phase does not appear within the period of observation; the crest temperature remains constant throughout the night. For most of the nights, a smooth transition between the two modes of cooling occurs at about midnight. A feature conspicuous by its absence is a "warm altitude belt," that is, an intermediate altitude at which a persistent temperature maximum is found on the slope. Local temperature maxima do occur for many of the nights, but they are transient and appear to be frequent at no one altitude level on the slope.

The "warm belt" or "altitude of maximum temperature" mentioned above has been observed in many studies, and the available literature is discussed in detail by Geiger (1957). One notes that the "warm altitude" belt tends to be regarded as a two-dimensional feature: some particular altitude on the slope. Regarded as such, it becomes a rather mysterious phenomenon. If the three-dimensional character of the cold airflow on most mountains is considered, however, such an altitude zone becomes possible as a consequence of the way in which a particular geometrical altitude transect is drawn relative to the estimated streamline pattern. This point was apparently first noted by Hayes (1941). Thus, a line (AB) can be drawn on figure 1A with a continuous increase in altitude which would probably show a definite "warm belt." This "warm belt" would reflect the fact that the air on the upper portion of the transect beyond point (C) would have been in contact with the surface heat sink for a much longer period of time than the air below (C); in other words, they originate at different intake zones. Note also that the importance of divergence centers such as (C) would not be apparent from a two-dimensional cross section of a slope. While they may appear as regions in the two-dimensional cross section of low slope, or "shelves," such shelves could be zones of convergence or divergence, depending on the sense and magnitude of the height contour curvature.

Short-period pulsing in the temperature field is found in the observations, particularly for August 22 and September 8, 1962, and September 16, 1964 (fig. 3). For August 22 and September 8, which are more clearly defined, pulse amplitude appears to decrease downslope, yet there is no detectable phase shift or change in period. The effect seems to be associated with the shift in the cooling pattern mentioned above. Since the nights of September 9, 1964, and June 12, 1966, began with an isothermal hillside, it cannot be said that pulsing is associated with a large lapse rate outside the cooled layer on the slope, as might be inferred from the 1962 data. Also, the position of maximum pulse amplitude on the hillside appears to vary between the nights.

Radiation

As shown by some direct measurements in 1964, (Bergen 1969), the net radiation above the canopy at a point on the slope averaged over the night is fairly well approximated by the average difference between the sky radiation measured at the base of the slope and the black body emission, which corresponds to the mean air temperature measured in the canopy. Assuming that such an estimate may be used for 2-hour periods, the net radiation for those intervals was estimated from the radiometer readings and the associated averages of the measured air temperatures in regard to time, height above the slope, and position along the transect, for the 5 nights of 1966 (table 2).

Sky radiation fell steadily through all the nights with the exception of June 16. There is more variation in the pattern of computed net radiation losses, including uniformly increasing values for June 11 and minima for various hours in the remaining nights. Comparison with the 1966 plots (fig. 3) indicates no clear relation between the time of such minima and the transition between the two modes of cooling mentioned above.

Variation of Potential Temperature along the Transect

The temperature drop along the hillside transect does not directly indicate the decrease in enthalpy of the air moving down the slope. Due to com-

Table 2.--Two-hour averages of sky radiation and estimated net radiation loss, 1966

Date	Time LST	°C. $\overline{T}m$	Ly $Min.^{-1}$ R_B	Ly $Min.^{-1}$ Rsk	Ly $Min.^{-1}$ R_o
June 11	2000	8.6	0.404	0.404	0.110
	2200	6.9	.374	.374	.128
	2400	2.0	.335	.335	.132
	0200	- .2	.321	.321	.132
June 12	2000	5.8	.494	.412	.082
	2200	5.4	.491	.385	.106
	2400	3.4	.477	.342	.135
	0200	.7	.459	.319	.140
June 14	2000	9.7	.522	.379	.143
	2200	8.8	.515	.383	.132
	2400	8.4	.512	.377	.135
	0200	6.7	.497	.340	.157
June 15	2000	8.3	.512	.399	.113
	2200	7.9	.509	.442	.067
	2400	6.4	.498	.395	.103
	0200	5.0	.488	.387	.101
June 16	2000	8.6	.498	.391	.107
	2200	6.4	.498	.419	.079
	2400	5.2	.489	.392	.097
	0200	6.2	.497	.398	.099

pressibility, downward movement of air along an isothermal hillside would correspond to an appreciable energy loss, the amount depending on the change in altitude and thus the mean atmospheric pressure from point to point on the flow.

"Potential temperature" directly reflects the enthalpy losses of the air on the slope. The potential temperature may be defined as the final temperature of a unit mass of air which is compressed to a pressure of 1,000 mb. without energy loss or gain, and without a change of phase in the water contained in the air.

The use of 1,000 mb. as the reference pressure is arbitrary, and in the case of the experimental hillside, it is more convenient to use the standard atmosphere pressure at point (C). Thus, since (C) is about 90 m. above the tower station, we would subtract 0.9° C.—90 m. times the adiabatic lapse rate of 0.98° C. per 100 m.—from the temperature at the tower station to obtain the potential temperature Θ.

When this transformation is applied to the base average temperatures ($\overline{T}$) as previously calculated we obtain a corresponding average potential temperature $\overline{\Theta}$, which is plotted (fig. 4) for 2-hour intervals for each of the nights against the downslope distance (X) measured along the streamline of figure 1A from point (C).

The potential temperatures plotted in figure 4 correspond to 2-hour averages of ($\overline{T}$) and show a more regular trend, in particular the temperature drop is close to linear on the slope itself, that is, for X less than 400 m. The pulses evident in the previous figures are almost totally smoothed out.

The two patterns of cooling are more evident in figure 4 than in figure 3, as is the significance of point (C), particularly on June 12 and June 15. A linear extrapolation on these nights would yield an almost constant temperature at (C), although the temperature at the base of the slope drops as much as 10° C. through the night. Temperature at point (C), then, is a function of the potential temperature field above the cooled layer on the hillside, while the potential temperature drop along the streamline depends primarily on the net radiation loss along the hillside.

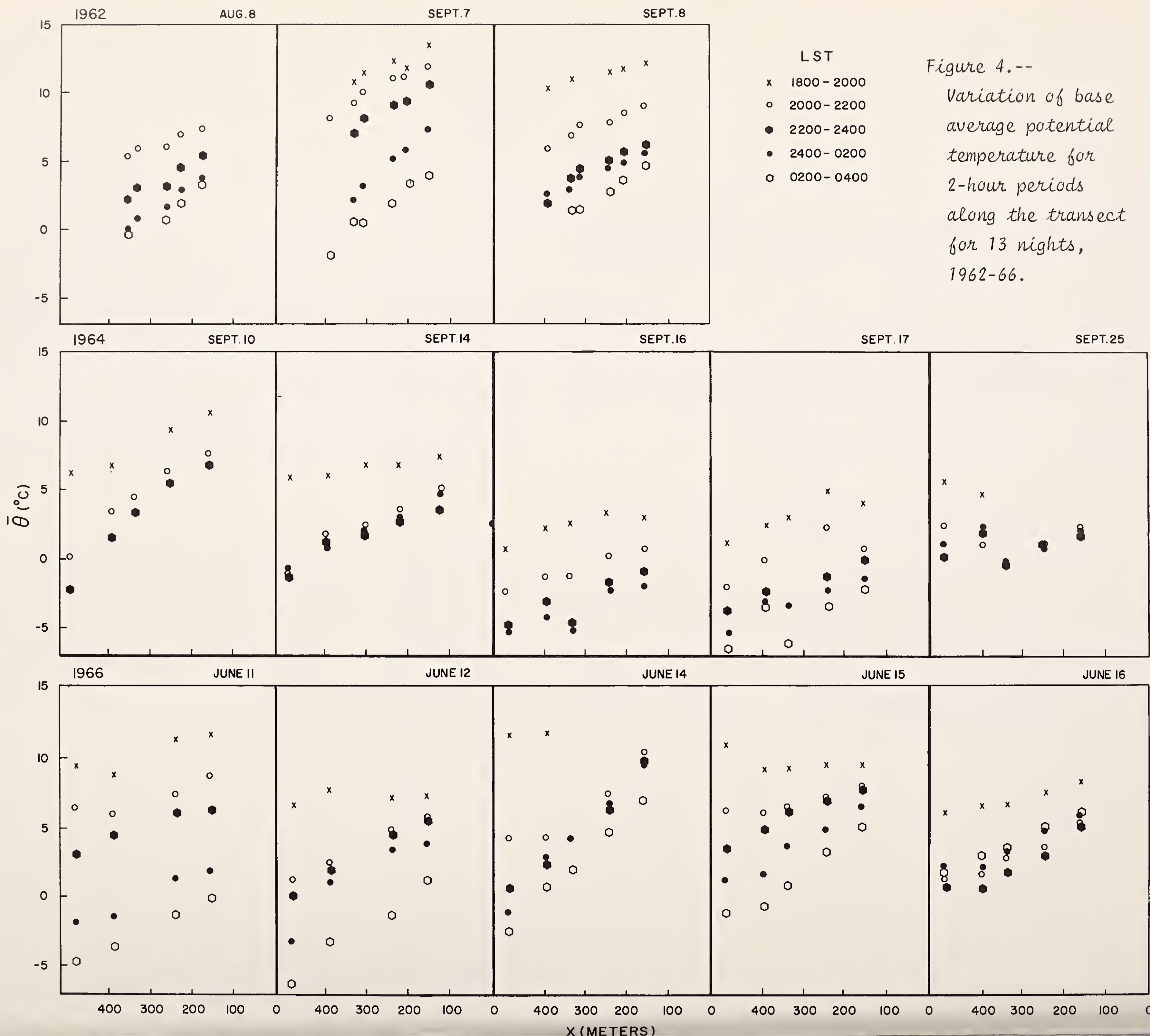


Figure 4.--Variation of base average potential temperature for 2-hour periods along the transect for 13 nights, 1962-66.

The data presented in figure 4 have a direct bearing on an assumption basic to many analytical solutions for the motion of cold air on a hillside. Most of these solutions involve what may be called a "perturbation assumption" in regard to the sensible heat advected along the slope by the drainage flow. The temperature drop along the slope is assumed to be small in relation to the drop in temperature with height in the air above the slope. The most extreme assumption is that the slope is uniformly cooled, which is the case first treated by Prandtl (Sutton 1953).

This assumption is quite distinct from the basic "pressure perturbation" assumption: that the pressure change on the slope is small relative to the total atmospheric pressure, which is easily defended and was in fact made above in evaluating the potential temperatures plotted in figure 4. It may be seen that the downslope gradient for either the potential temperature or the temperature field itself is at least the order of 2 to 10 times that on the hillside at the onset of cooling, and far greater than would be anticipated for the free atmosphere over the valley at sunset.

Composite Potential Deficit Profiles

There are many points in common between the flow of cold air from a hillside and free convection phenomena observed in the laboratory, such as the ascent of air over a vertical heated plate. Such laboratory flows are often found to exhibit "similarity" in both the velocity and temperature distribution.

The notion of similarity as applied to the case of cold air drainage off such a slope and referring to the potential temperatures on the slope is equivalent to assuming that some characteristic potential temperature (Θ_s) and some length (L_s) exist, which are functions only of downslope distance (X) from the origin of the flow measured along the streamline, and are such that ($\Theta - \Theta_o$) at any level above the slope (Z) conforms to the relation:

$$\theta - \theta_0 = (\theta_s - \theta_0)\, f(Z/L_s)$$

where Θ_c is the potential temperature at point (C).

While many choices are possible for Θ_s and L_s, the average potential temperature below the first inversion above the slope ($\overline{\Theta}$) discussed above must be a legitimate scaling temperature if such exists. The most accessible length is the height of the first inversion above the slope, which has been denoted as (L) above.

While such scaling would appear sufficient for a bare slope, the temperature measurements discussed were made in a random array of trees. The streamlines (fig. 1A) must have relevance to the actual movement of air only as a spatial average over lengths of the order 20 m. or so. The situation is in many respects analogous to flow in a porous medium, with the spaces between trees on the slope being analogous to the pore spaces. Since the radiating foliage is the primary heat sink for at least the portion of the airflow above the first inversion, (Z = L), there must be a temperature drop from the center of these spaces to the foliage surface of the trees. Unfortunately, foliage temperatures were not measured directly during the observation. The 1962 measurements do show that the air temperature measured between trees does not vary appreciably between stations some 20 m. apart at the same downslope distance from the origin; the average difference between such stations was less than 0.3° C. Thus, the approximate linearity of the potential temperature $\overline{\Theta}$ shown on the transects (fig. 4) is apparently more than an artifact of the choice of stations.

The apparent height scales for June 14, 15, and 16 are 11m., 7 m., and 10 m. for the stations I, III, and the tower station, respectively. The scale for station IV was taken as 9 m., an intermediate value between that for station III and the tower station and in accord with the relative position of the inversion at station IV found in the 1964 observations.

The apparent height scale (L) for the nights of June 11 and 12 is less by about 15 percent—about 6 m. at the midslope station. The scale was assumed to be less by the same proportion at the remaining stations, although the variation at these less densely instrumented profiles was not as clear.

The number of data levels at each station varied between observation periods, with most of the levels above (Z = L) inoperative during the first night, June 11.

Even at the densely instrumented stations at the tower and at midslope, (L) could not be resolved to within less than a meter; the estimates for stations I and IV are thus even more conjectural. The temperature (θ_o) at the point (C) was estimated by linear extrapolation from the values of $\bar{\theta}$ at the stations on the slope (fig. 4).

The results of a composite plot of (Z/L) versus

$$\frac{\theta - \theta_0}{\bar{\theta} - \theta_0}$$

for the 1966 observations are shown in figure 5 for all stations. The plots are for 2-hour average temperatures, with the individual station's data coded in the legend.

The hypothesis of similarity as stated in the equation above would require that these profiles for the individual stations superimpose.

The plots suggest this is the case for most of the intervals of measurement. In evaluating the superposition, it must be noted that the nominal accuracy of these temperature measurements is about ±0.5° C. at all stations on the slope. Since θ decreases appreciably down the transect, this would imply a much larger relative error for station I than for the tower station.

Below the first inversion, superposition is within 10 percent at all levels between the stations. Above the first inversion, the superposition is of more variable quality, but still within 20 percent.

To compare the composite profiles for different intervals of observation on the same night and between nights, a pair of polynomial segments were fitted to the profile for the last three profiles of June 15. These segments are shown by the dashed lines on the plots for all the observation intervals of June 12, 14, 15, and 16. For most of the intervals, the deviation from this fitted profile is nonsystematic and amounts to not more than 20 percent.

For the observations of June 11 and 12, the reduction of scale length appears to be associated with a relative expansion of the warm layer above the inversion. While the deficiency of the upper-level data for June 11 makes comparison with the profile fitted to the June 12 observations doubtful, the superposition appears satisfactory below (Z= L).

The approximate validity of the similarity assumption is more than a curiosity, and can serve as a valuable tool in the quantitative analysis of the temperature regime on the slope, as shown by Katkov (1965). The polynomial segments that appear to yield a plausible fit to the scaled potential temperature deficit profiles are of low order: a parabola below (Z/L = 1), and a cubic equation above that level with the physically plausible boundary conditions of a vanishing gradient at (Z = L), and zero deficit at (Z = 2.3L).

Conclusions

The nocturnal temperature distribution on nonuniform slopes reflects the three-dimensional movement of air off these slopes, both in the spatial variation of the average temperatures of the active layer of air, and in the vertical variation of air temperature through that layer. In such situations, the geometrical summit of a slope may be less relevant than the centers of divergence for flow on the slope computed on the assumptions of continuity and that the flow is normal to the elevation contours.

On the simple slope studied, the potential temperature profiles along an estimated streamline show approximate similarity.

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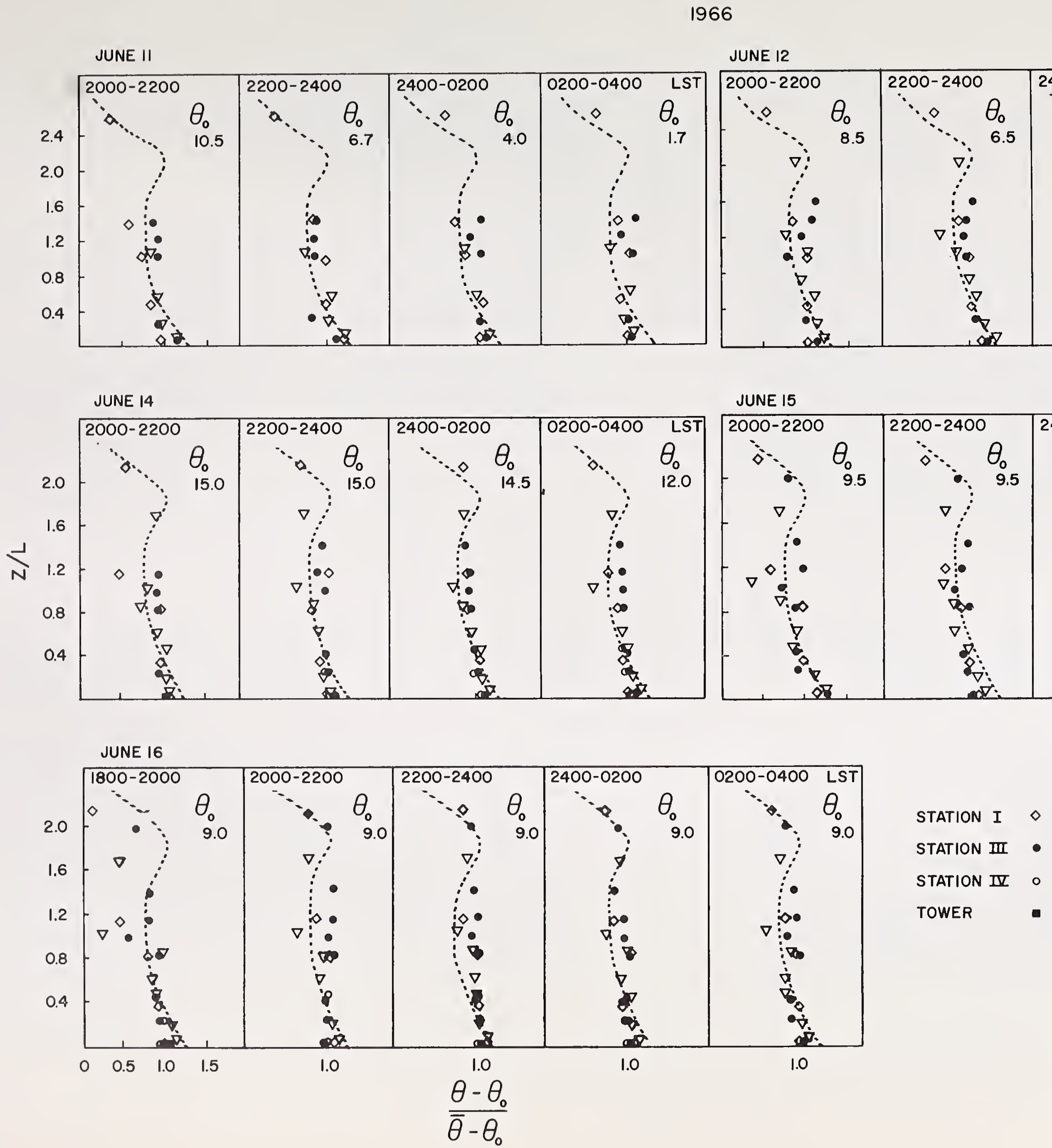

Figure 5.--
Composite profiles of potential temperature deficit, June 1966.

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Profiles of air temperature on a forested mountain slope show an inversion below the top of the tree canopy. The slope tends to cool with constant downslope temperature gradient in the early part of the night; gradients increase in the later hours. Plots of potential temperature indicate, for some nights, a point of almost constant temperature halfway up the hillside. This point can be identified with a center of divergence for cold air moving off the slope. The potential temperature deficit relative to the temperature at this point shows approximate similarity between vertical profiles at the different stations when the profiles are scaled by the height of the first inversion above the slope and the average potential temperature deficit below that inversion.

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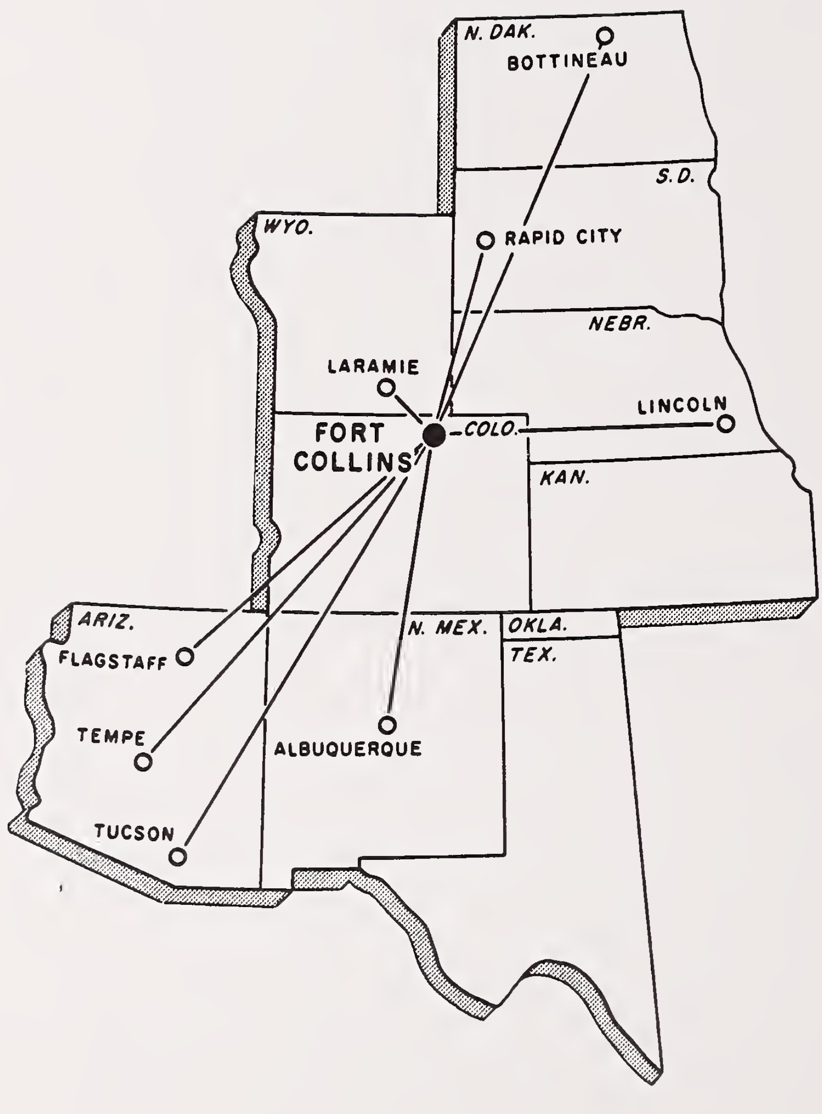
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